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Comment on: "Petrography of Snap Lake kimberlite dyke (Northwest Territories, Canada) and its interaction with country rock granitoids"

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Abstract

Fulop et al. (2018) infer that the Snap Lake kimberlite intrusion (NWT Canada) was emplaced as a single batch of magma and that internal variations are a consequence of alteration and interactions with granitoid host-rocks by deuteric fluids. However, field, petrological, geochemical and theoretical constraints support multiple magma batches, and show that the deuteric alteration hypothesis is implausible. Rather we propose that alteration is primarily the consequence of reactions between the kimberlite, country rock xenoliths and surrounding granite with groundwaters during cooling and serpentinisation at temperatures between 400°C and ambient.

Introduction

Fulop et al. (2018) present new data on the Snap Lake kimberlite intrusion in which they interpret variations in mineral assemblages and bulk rock geochemistry as the consequence of alteration by deuteric fluids combined with interaction between kimberlite magma with granite host rocks and xenoliths. They question the interpretation of previous studies (Field et al., 2009; Gernon et al., 2012) that there are marked compositional variations of igneous origin in the kimberlite magmas leading to primary zoning within the intrusion. While these rocks are indeed profoundly altered, the main new interpretations require implausible amounts of volcanic fluids to be transported with the kimberlite magma. Further, this hypothesis is not consistent with the observed alteration mineral assemblages, which we show here were formed at low temperature by very water-rich fluids. Observations at Snap Lake can be largely explained by compositional variations in the kimberlite overprinted by alteration (predominantly serpentisation) by groundwater. However, some features of these rocks could have originated by reactions between melts, fluids, primary igneous crystals and entrained xenoliths during the initial stages of solidification.

Fulop et al. (2018) discredited our earlier studies making the false statement that Gernon et al. (2012) "mapped [rock types] from underground photographs". However, field data and observations were collected underground at Snap Lake with De Beers Canada geologists over several months using well-established underground mapping and direct observations of rock faces. High-resolution oriented photographs of rock faces were used only for image analysis of key parts of the dyke (see Field et al., 2009; Gernon et al., 2012). As the underground mine workings were flooded after closure of the Snap Lake Mine in 2015, we revisit data and detailed field and petrographic observations that formed the basis of the studies by Field et al. (2009), Gernon et al. (2012), and Ogilvie-Harris (2012).

In this comment, we will develop three topics: (1) the implausibility of magmatic fluids being the main cause of alteration; (2) the evidence that variations across the intrusion have a magmatic signature and cannot be explained entirely by alteration; and (3) a model of largely low temperature alteration caused by reaction of the fluids generated by serpentisation with acidic cratonic groundwaters and brines, which overprints the primary variations related to igneous processes.

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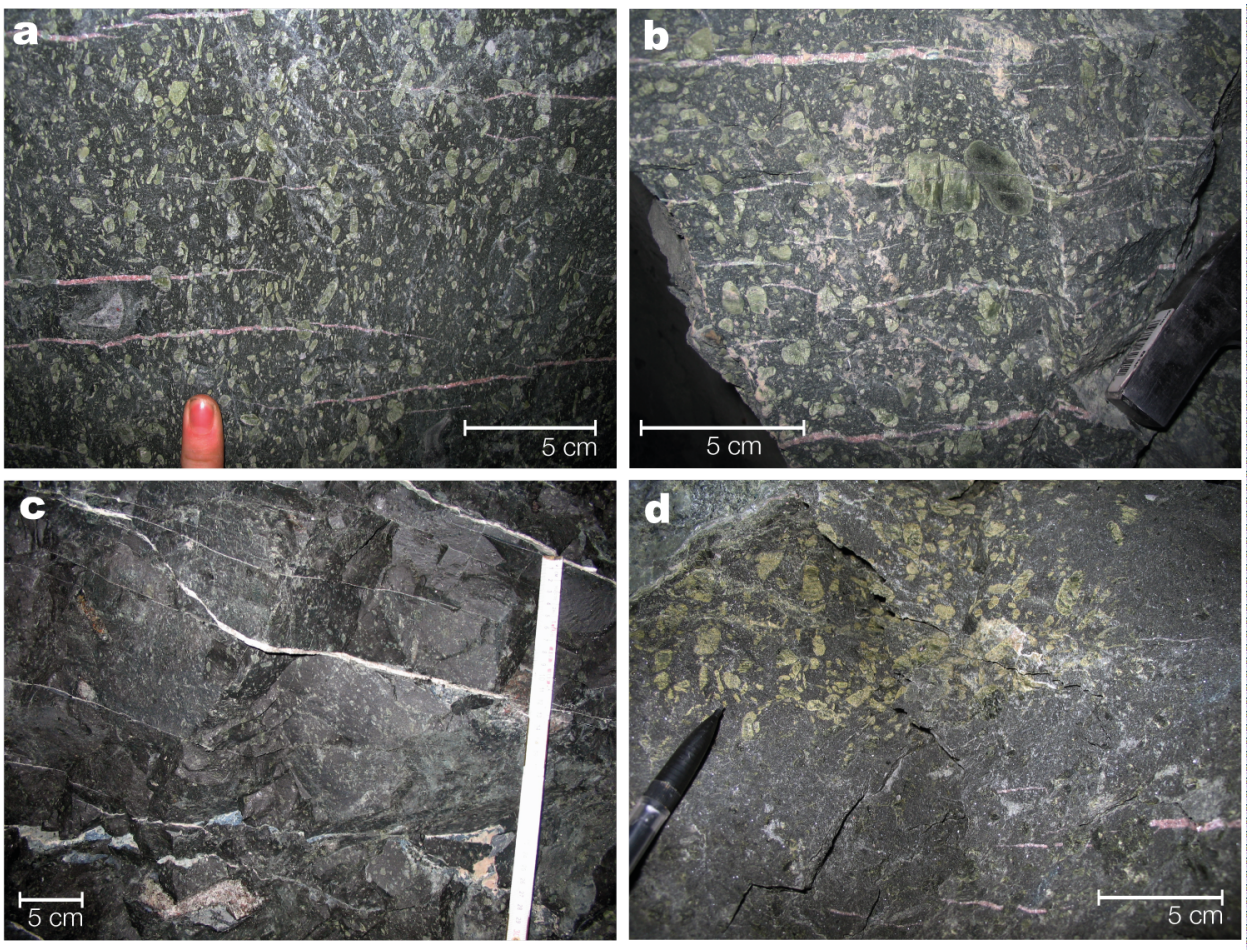

Figure 1: Field photographs showing some characteristic features of Snap Lake kimberlites. (**a**) Olivine-rich macrocrystic kimberlite (ORK) showing a strong sub-vertical fabric defined by the alignment of elongate olivine macrocrysts. Note the dyke margins are roughly parallel to the calcite veins (i.e. sub-horizontal), so the fabric is developed perpendicular to the dyke margins. (**b**) ORK showing a high concentration of large olivine macrocrysts. (**c**) Kimberlite transitional between ORK and olivine-poor kimberlite (OPK), showing a patchy distribution, and relatively low abundance, of olivine macrocrysts. The scattered distribution of olivine indicates that the low olivine abundance is a primary feature, inconsistent with a deuteric alteration overprint (i.e. there is no reason why deuteric alteration would selectively pseudomorph some olivine macrocrysts, and completely evade adjacent olivine macrocrysts). (**d**) Cognate xenolith of ORK in OPK, showing sharply-defined contacts (Field et al., 2009). Note that olivine macrocrysts are visible in the OPK, but less abundant. Such features are hard to explain in the context of deuteric alteration, which invokes a gradational transition between ORK and OPK (Fulop et al., 2018). In addition, a deuteric alteration overprint of OPK would need to selectively exclude many olivine macrocrysts, which remain intact (Figs. 2 and 3).

Alteration—fluid sources and conditions

To be absolutely clear we are not claiming that there was no deuteric alteration. Following and during dyke emplacement there will have been a brief period of cooling and crystallisation during which there could have been reactions between the dyke's constituents (primary crystals, melts, magmatic fluids and entrained xenoliths). However, here we show that any reactive fluids generated at this stage cannot be the main source of fluid that caused the predominant alteration.

The presence of igneous phlogopite in the Snap Lake intrusion indicates that the kimberlite was wet, so the central question is whether the amount of water transported by the kimberlite into the location of emplacement could account for the alteration. The main alteration products (serpentine and phyllosilicates) require a huge amount of water to have been transported. In order to constrain the amount of water it is necessary to know the depth of emplacement, the solubility of water in the melt and the porosity of the magma developed during and perhaps shortly after emplacement. Here we estimate an upper bound on this amount.

The depth of emplacement is well constrained by (U–Th)/He age dating of zircons and apatites (Flowers et al., 2006) which demonstrate very long-lived stability of the Canadian shield with no more than 1.5–2 km of erosion from mid-Proterozoic times. Thus we use a lithostatic pressure of 20 MPa for a maximum depth of 2 km plus 10 MPa for the typical overpressures associated with dyke emplacement for our calculations. The solubility of water in kimberlite melts is constrained by experimental studies (Brooker et al., 2011). **Silica undersaturated melts have solubilities similar to basalt [Richard's thoughts]**, while there is some evidence that water is more soluble in carbonate melts (Keppler, 2003). Based on the Snap Lake mineralogy (olivine, monticellite, phlogopite, oxides and minor carbonate) we assume a dissolved water content of 2.0 wt% using experimentally constrained solubility laws. Using a magma temperature of 800°C constrained by low-pressure phlogopite stability (Righter and Carmichael, 1996), the density of water at 30 MPa is about 50 kg/m^3. Assuming a conservative magma porosity of 10% and a magma density (crystals plus melt) of 2750 kg/m^3, then 0.2 wt% could be transported as an exsolved phase. There is however a trade off in that the higher the porosity the lower the melt fraction so the total exsolved and dissolved water will decrease with increasing porosity. Phlogopite typically contains about 2.5 to 3 wt% water and it seems reasonable to suppose that the melt from which it crystallised contained comparable or greater water contents. The mode of phlogopite in some Snap Lake rocks exceeds 50% suggesting a melt water content of at least 1.5%, consistent with the above calculations. This water content is similar to that estimated by Soltys et al. (2018) for the Bultfontein Group I kimberlites (2.1-2.2 wt%).

Now consider the water budget needed for alteration. Table 6 in Fulop et al. (2018) shows typical loss on ignition values of around 15 wt%. Given that much of the Snap Lake magma is highly porphyritic the water would need to be carried in the melt, so the requirement is for the magma to transport 25–30 wt% water—more than an order of magnitude greater than what is plausible for magmatic water content. We conclude that another source for almost all the water is required. Given that the granite country rocks contain only minor amounts of water (0.1 to 0.4 wt% according to the modal data), the ground water is the only plausible source and we therefore dismiss the hypothesis that the alteration is largely deuteric (i.e. involving magmatic water).

Fulop et al. (2018) also invoke magmatic CO_2 in their reactions with formation of dolomite as a by-product of serpentisation. It is hard to distinguish primary from secondary carbonates in such highly altered rocks, and we concur that primary kimberlite magmas are commonly carbonatitic in character. Thus magmatic CO_2 might have played a role, but for the same reasons as argued for water, exsolved CO_2 cannot have supplied much. Alternatively a decarbonation reaction of igneous carbonate may have occurred at magmatic temperatures to release magmatic CO_2. However, there is a simpler alternative explanation for the presence of abundant carbonate (notably in late stage veins, Fig. 1) at Snap Lake that derives from the extensive literature on low temperature serpentinisation (e.g. Charlou et al., 2002; Oze and Sharma, 2007). Formation of carbonate is ubiquitous where very highly alkaline and highly reduced serpentising fluids mix with neutral or acid external water (Palandri and Reed, 2004; Schrenk et al., 2013), often moderated by microbial activity. Below we will develop an interpretation in which serpentinisation generates fluids which react with groundwater at the intrusion margins and in the near-field host rock. A key point developed below is that the alteration mineral assemblages do not support involvement of a CO_2-rich fluid.

One further difficulty for the deuteric hypothesis is that in a low angle (15°) dyke (almost a sill) the fluids escaping from the magma should move upwards. Thus one might expect to observe much stronger alteration effects in the roof of the intrusion than the floor. Such asymmetric effects are not reported in this and previ-

ous studies of Snap Lake. Our detailed observations of actual sections (Field et al., 2009; Gernon et al., 2012) agree with the schematic sections (Figures 2 and 3) in Fulop et al. (2018) that show the variations across the intrusion are symmetric. One possibility for generating symmetrical sections is if the magma intrusion had high fluid (gas) overpressures resulting in decompression and outflow of volcanic fluids. However, we have already shown that the amount released is far too little to cause the proposed alteration.

Origin of mineralogical and geochemical variations

The hypotheses that variations within the Snap Lake intrusion are the consequences of diverse magma compositions, of interaction between these magma and country rock, and alteration are not mutually exclusive. However, Fulop et al. (2018) adopt the position that the latter two processes can explain all the observations. Our intention here is to show unequivocally that there are primary magmatic variations, reiterating some of the key observations already presented in Field et al. (2009) and Gernon et al. (2012), and drawing on some of the new data presented by Fulop et al. (2018).

Field observations and lithofacies variations

A key assertion by Fulop et al. (2018) is that the Snap Lake intrusion is comprised of a single magma and the variations reflect differences in alteration between the margins and centre, with replacement of abundant olivine by phlogopite. However, there are several marked differences in the characteristics of the marginal and central rocks. First there is a significant difference in olivine size and abundance clearly evident in Figures 4 and 5 of Fulop et al. (2018) as well as in the data presented by Gernon et al. (2012).

Figure 1 shows contrasted macroscopic textures observed in ORK and OPK. Even where differential alteration has occurred (for example, as evidenced by changes in olivine colour), olivine macrocrysts are discernible, and their abundance varies locally to define olivine-rich and olivine-poor regions (Fig. 1). In some areas, a uniformly scattered distribution of olivine is observed (Fig. 1c), which would not be expected if the intergranular kimberlite formed as a result of deuteric alteration. The ORK crystals show a strong sub-vertical fabric (Fig. 1a) that is not developed in the OPK (Figures 3 and 4 in Gernon et al., 2012), pointing to fundamental differences in magmatic emplacement for these lithofacies.

Fulop et al. (2018) argue that the changes from HK1 (i.e. defined as ORK in our papers) to HK6 (i.e. OPK

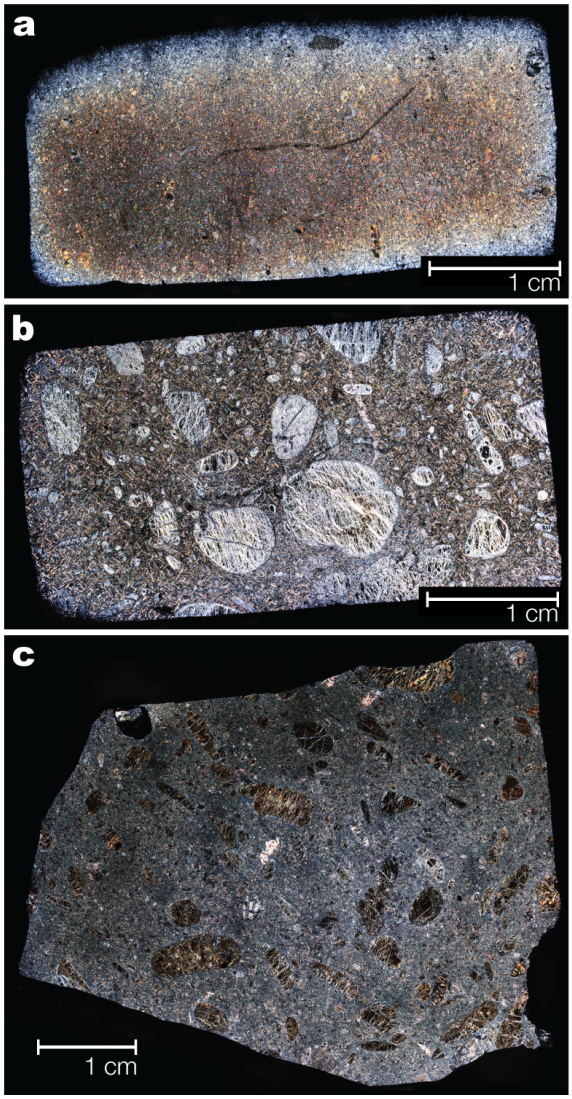

Figure 2: Photomicrographs (transmitted light) of three main kimberlite lithofacies identified in Field et al. (2009) and Gernon et al. (2012). (**a**) Olivine-poor kimberlite (OPK). Olivine macrocrysts are heavily serpentinised yet still visible, and sparse relative to the ORK. The deuteric alteration hypothesis of Fulop et al. (2018) requires that the abundance of olivine macrocrysts was originally equivalent to the ORK [see (c)], and these crystals have since been pseudomorphed by phyllosilicates. Small serpentinised olivines are visible, but the purported pseudomorphs are not visible (see also Fig. 3). (**b**) Transitional ORK–OPK variety studied by Field et al. (2009). Again, serpentinised olivine macrocrysts are clearly discernible. Note the abundance of platy phlogopites in the groundmass defining a decussate-like texture. (**c**) Olivine-rich kimberlite from sample S1 (Gernon et al., 2012). Here, the modal abundance of olivine macrocrysts is much higher than that of the OPK, and there is significantly less phlogopite. Note the slightly smaller scale in (c).

in our papers) are always gradational. However, Fulop et al. (2018) only provide photographs of discrete core samples (rather than outcrop-scale relationships), making the evidence for a gradual transition unclear to the reader. Field et al. (2009) and Gernon et al. (2012) show that abrupt changes between varieties are common (e.g. Fig. 1d), and occur at all levels in the dyke. Both gradual and abrupt changes can occur during magma mingling, so where the variations are gradual is not convincing evidence for an alteration model (but alteration might well lead to more diffuse contacts).

Olivine and phlogopite textures

Fulop et al. (2018) present textures from which they infer replacement of olivine macrocrysts by phlogopite and other phyllosilicates. We do not find them convincing; their approach of drawing a bold red outline obscures the supposed boundary of the pseudomorph. Rather, textures of the OPK, observed under Scanning Electron Microscope (SEM), are inconsistent with a deuteric alteration overprint as now demonstrated. The images (Fig. 3a–3e) are taken from OPK at the top of the profiles ND1-01 and ND2-06 shown in Field et al. (2009) and Gernon et al. (2012). First, olivines are present as discrete macrocrysts, with little evidence for phyllosilicate replacement or substitution pseudomorphing (Fig. 3). Second, there are many quite abrupt contacts between phlogopite and olivine (Fig. 3b–3d), and the phlogopite laths are randomly orientated (Fig. 3a–3d), reminiscent of decussate textures. Fulop et al. (2018) accept that the 'random, cross-cutting orientations' of phlogopite evidence late-stage crystallisation from a magma (*sensu* Gernon et al., 2012). Third, serpentinised monticellite is found poikitically enclosed in apatite (Fig. 3e). The apatite clusters appear to grow up to the edges of (and partially enclose) the phyllosilicates (Fig. 3e), suggesting the latter are not late replacement features. Fourth, the overall distribution of phyllosilicates (e.g. ND2-06a, Fig. 3a) is not at all reminiscent of the textures and crystal size distribution defined by olivine in ORK (e.g. Fig. 3f), which Fulop et al. (2018) argue to be the OPK 'protolith'. Finally, many phlogopites exhibit lath-like geometries and are surrounded by serpentine groundmass, with the latter associated with other primary kimberlite minerals such as spinel and titanite. Such textures cannot be generated by overprinting olivine macrocrysts and microcrysts.

Diamond and olivine distributions

The diamond and olivine grain sizes are positively correlated (Field et al., 2009), which is also an indicator of variation in the characteristics of the magmas. Field et al. (2009) recovered bulk samples from the ORK and OPK lithofacies (Gernon et al., 2012), ranging from 30–120 tonnes (averaging 60 tonnes) in mass. Sampling on this scale is essential to make a representative assessment of diamond grade (Rombouts, 2003) and distribution across a range of diamond sizes. Ore extraction during bulk sampling was very carefully supervised to mitigate contamination of different lithofacies. Field et al. (2009) show clear differences in the diamond and olivine distributions for OPK and ORK (Fig. 4). They also show that the cumulative average diamond size varies laterally in the intrusion and are strongly correlated with lateral lithofacies variations. The cumulative average diamond size is lowest in areas that are dominated by OPK (Fig. 9, Field et al., 2009), and highest in ORK-dominated regions of the dyke. These observations are explained by primary magmatic variations.

Fulop et al. (2018) present microdiamond data showing the different lithofacies are almost identical (Fig. 17 in their paper), but do not mention that the larger diamonds show significant and systematic variations (Field et al., 2009). This discrepancy could be explained by the different type of analysis (microdiamond as opposed to macrodiamond), widespread mixed sampling strategy (i.e. combining samples from different locations, leading to mixing of lithofacies; see Fig. 1 in their paper), and relatively small sample size (175×8 kg samples). Fulop et al. (2018) show bundled lithofacies (i.e. HK1–2 and HK3–6), which unfortunately makes it difficult to draw meaningful comparisons between the datasets presented in Fulop et al. (2018) and Field et al. (2009).

Phlogopite chemistry

The marginal rocks are richer in phlogopite and poorer in serpentinised olivine from modal data and bulk geochemistry (higher K and lower MgO) and occupying distinctive fields in Figure 15 of Fulop et al. (2018). In contrast to Fulop et al. (2018), we found phlogopite-rich margins in the area with metavolcanic country rock (see Figure 3 in Gernon et al., 2012). We interpret these relationships as due to a variation in melt compositions from which the phlogopite precipitated.

We find clear differences between the OPK and ORK phenocrystal (matrix) phlogopites reflected in their TiO_2 and Cr_2O_3 contents (Fig. 5). ORK phlogopite ranges between 0.85 and 2.2 wt% TiO_2, whereas the ORK phlogopites have a lower range between 0.10 and 0.85 wt% TiO_2 (Fig. 5). The Cr_2O_3 are very low in the ORK (0–0.2 wt%) and show a considerably wider range in the OPK (0–2 wt%). Similarly, Kopylova et al. (2010) showed the presence of low (0.1–0.5 wt%) and

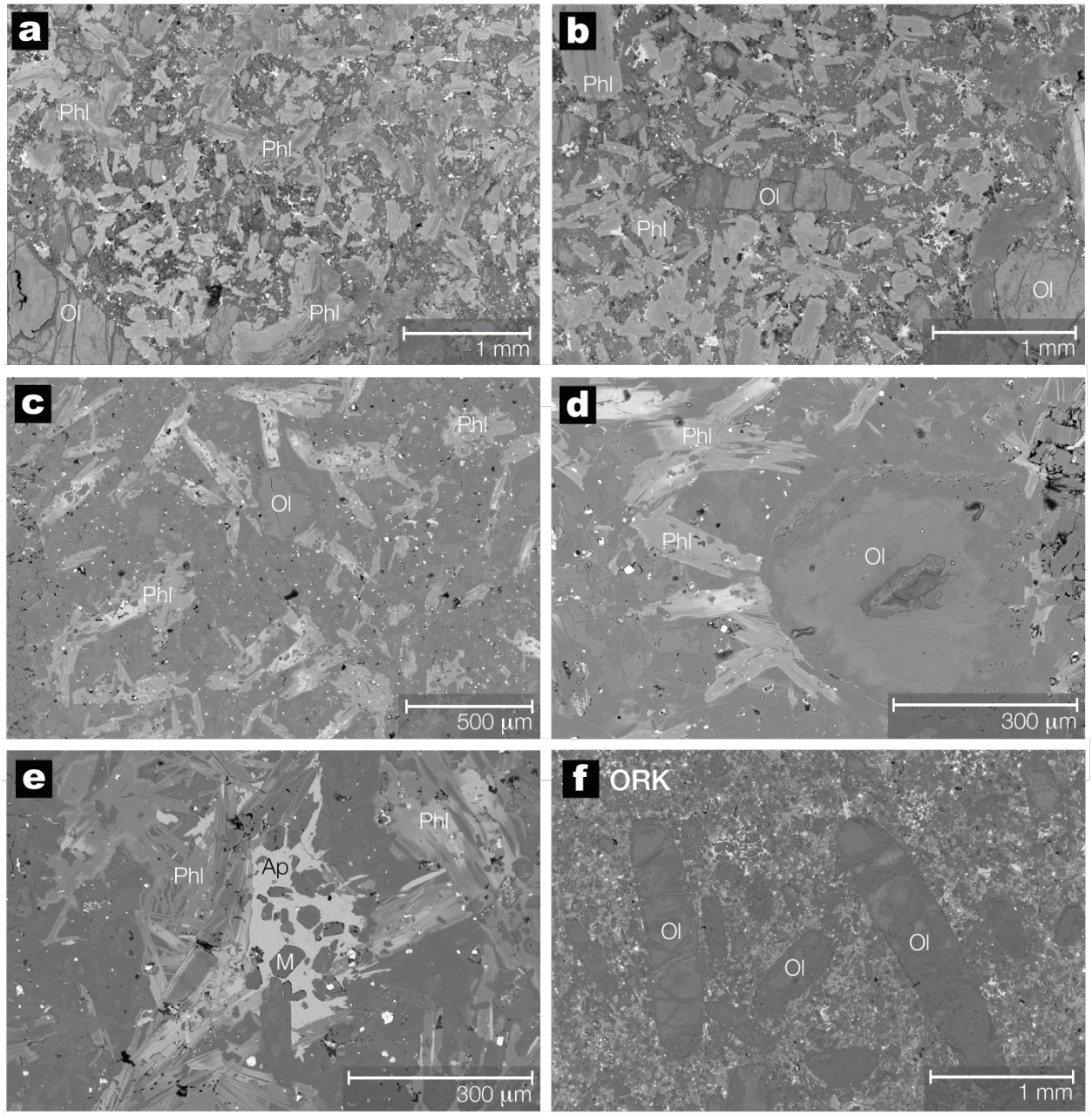

Figure 3: SEM photomicrographs of OPK from Snap Lake. (**a**) and (**b**) OPK from the top of profile ND2-06 (Field et al., 2009; Gernon et al., 2012). Here, olivine occurs as discrete macrocrysts. The phlogopite laths are randomly orientated and do not appear to be pseudomorphing olivine macrocrysts [compare with ORK shown in (f)]. (**c**) OPK from the top of profile ND1-01, showing randomly oriented phlogopite laths, some of which abruptly terminate at olivine crystal boundaries. (**d**) OPK from the top of profile ND1-01; again the phlogopites do not pseudomorph olivine macrocrysts and rather terminate abruptly at their outer boundaries (i.e. the textures are inconsistent with pseudomorphing). (**e**) OPK from the top of profile ND1-01; here apatite poikilitically encloses serpentinised monticellite and adjacent phlogopites, indicating that these are early formed crystals. (**f**) For comparison, ORK showing a characteristic fabric defined by olivine macrocrysts (Gernon et al., 2012). Note that Phl: phlogopite; Ol: olivine; M: monticellite, and Ap: apatite.

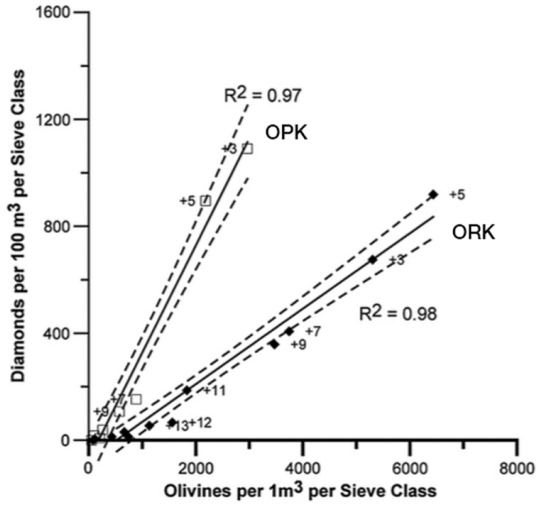

Figure 4: A comparison of the proportions of olivine and diamond retained for different DTC sieve classes for ORK (black diamonds) and OPK (white squares). The dashed lines signify 95% confidence intervals around fitted linear distributions (after Field et al., 2009). A linear correlation is evident for both lithofacies, which are distinct in terms of diamond distributions.

high (1–1.6 wt%) Cr_2O_3 phlogopites at Snap Lake, yet Fulop et al. (2018) did not measure Cr_2O_3 during phlogopite analysis (Supplementary File EST1 in their paper). Indeed Cr_2O_3 content is regarded as an important petrogenetic indicator in kimberlites (Reguir et al., 2009; Sobolev et al., 2009; Soltys et al., 2018). The textures shown by Fulop et al. (2018) (Figs. 11 b–c in their paper) are consistent with resorption and regrowth (cf. O'Brien et al., 1988), and typical of cases where magma recharge or convection through a temperature gradient have been proposed. Thus the simplest explanation for the observed differences in phlogopite chemistry is that OPK and ORK were formed from different magmas.

Problems with isocon analysis in kimberlite studies

The isocon method used by Fulop et al. (2018) is problematic in kimberlites because it requires (a) access to an unaltered sample, and (b) that the compositions of the rocks were originally the same (Grant, 1986; Guo et al., 2009). Fulop et al. (2018) assume that kimberlites in the centre of the dyke (i.e. HK1) are fresh, but these too are heavily serpentinised (Gernon et al., 2012). Using this method they also needed to assume that HK1 is the protolith of HK6, so circular reasoning is employed. There are further complications in using 'average fresh granitoid' (tonalite with K_2O < 2 wt%) as the granite protolith, because country rocks of the Defeat pluton suite (2.61–2.59 Ga) exposed in this region are highly heterogeneous (banded on a cm–m scale), comprising granodiorite, tonalite, and monzogranite with muscovite-bearing pegmatite locally (Stubley, 2000). Grant (1986) and Guo et al. (2009) stress that the isocon approach requires progressively altered samples and the starting material must be near identical for such normalisation methods to produce valid results.

Granitoid clasts and assimilation

Fulop et al. (2018) speculate that locally derived xenoliths can interact with magmas. They also suggest that granitic xenoliths are partially digested in the melt, contributing to local increases in Si and Al compositions in the kimberlite, and identify 'haloes' enriched in phlogopite around granitoid xenoliths (purportedly shown in Figures 5 and 9 of Fulop et al., 2018). We find that granitic clasts near the upper dyke contact are angular (i.e. not consistent with partial digestion), and directly in contact with olivines, with negligible phyllosilicate enrichment (Fig. 6).

There are major difficulties for low temperature kimberlite magmas assimilating highly refractory granitic rocks. Although not accurately constrained, estimates of kimberlite magma temperatures are around 800°C at shallow crustal depths; experimental studies indicate that at low pressure (20 MPa) phlogopite will not be stable above 900°C (Wones and Eugster, 1965; Righter and Carmichael, 1996). Granites are, however, very refractory because of their low water contents. A very small amount of melt should be formed at <900°C due to biotite breakdown, but temperatures typically need to exceed 1000°C for significant melting of granite xenoliths (e.g. Tuttle and Bowen, 1958) and about 1100°C for tonalite. It is not thermodynamically feasible to digest granitoid clasts in a kimberlite melt at these shallow crustal depths.

A simpler explanation for elevated mineral concentrations around some lithic clasts are thermal processes (e.g. quenching around cold clasts), or fluid dynamic phenomena such as flow alignment, which is common in magmas with such high crystal concentrations (Smith, 2000).

Mineral stability

Finally, we consider the stability of alteration phases reported in the Snap Lake kimberlite. Figure 7 is a T–X_{CO_2} diagram at 50 MPa like one cited by Fulop et al. (2018), but calculated using a different thermodynamic dataset. The diagram shows that all the alteration assemblages and proposed reactions only occur at temperatures below 400°C and indeed can occur at much

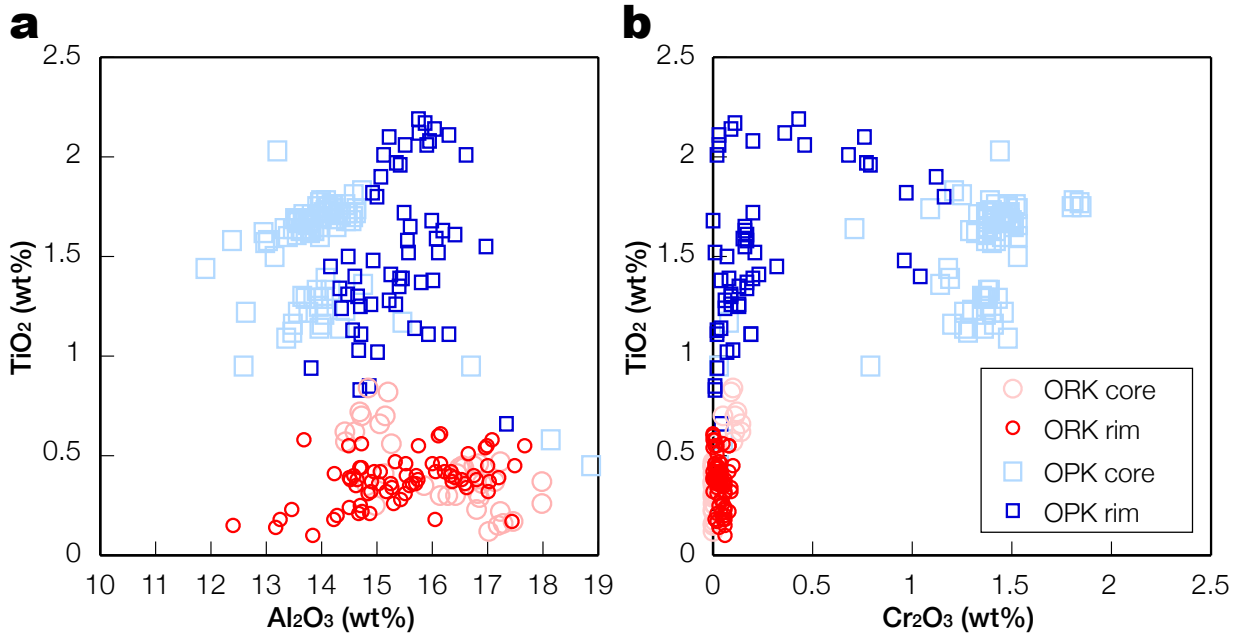

Figure 5: The defining chemical characteristic of OPK and ORK phlogopites and the general core-rim trends for TiO_2 plotted against (**a**) Al_2O_3; and (**b**) Cr_2O_3.

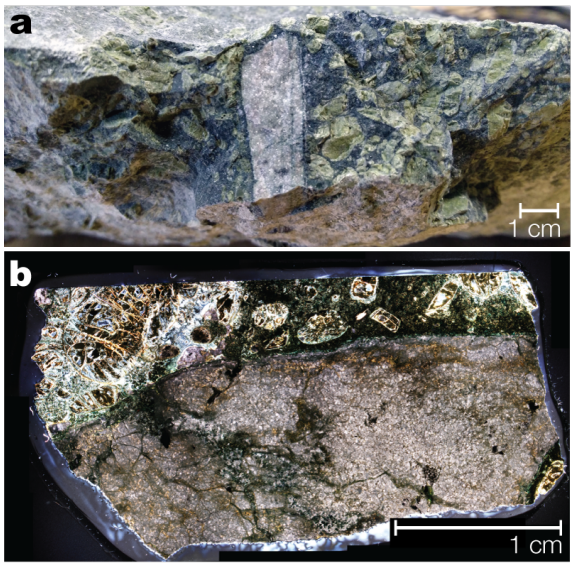

Figure 6: (**a**) Photograph of a granitoid lithic clast recovered near the upper dyke contact, showing sharp boundaries and local concentrations of olivine macrocrysts. (**b**) Thin section of a granite clast showing a sharp boundary with the adjacent kimberlite—a phyllosilicate-rich 'halo' is not visible.

lower temperatures approaching ambient. This figure excludes a deuteric origin for the assemblages. Fulop et al. (2018) suggest that the lack of perovskite and primary calcite in parts of the dyke is the result of deuteric alteration through the reaction:

$$2 \cdot Mg_2SiO_4 + CaCO_3 + CO_2 + 2 \cdot H_2O \longrightarrow Mg_3Si_2O_5(OH)_4 + CaMg(CO_2)_2 \quad (1)$$

Ca-depletion is attributed to this reaction, but this reaction causes Ca to be redistributed among minerals rather than lost from the system. Figure 7 shows that reaction 1 is encountered only for very water-rich fluids (c. 98.2–99.9% H_2O) and temperatures ranging from about 340–380°C. These temperatures mark the upper limit of dolomite + serpentine (liz), minerals either observed or reported from X-ray powder diffraction results (Tables 2 and 3, Fulop et al., 2018). The assemblage: serpentine (liz) + dolomite + talc (Tables 2 and 3; Fulop et al., 2018) is restricted to temperatures below about 340°C depending on fluid composition (Fig. 7). For essentially pure H_2O fluids both these assemblages could be stable at temperatures less than 150°C. Thus, we infer that alteration of the dyke resulted from influx and interaction with ground water during and after cooling.

Geochemical modification of the adjacent granites described by Fulop et al. (2018) can also be explained

by relatively low temperature alteration by serpentinising fluids. In other geological contexts, alteration of country rocks in contact with serpentinites results in rodingitization, which in granites can involve local Ca- and Mg-metasomatism and result in Si, K and Na depletion (Wares and Martin, 1980; Normand and Williams-Jones, 2007). There is no necessity to invoke higher temperature magmatic interactions to explain the observations.

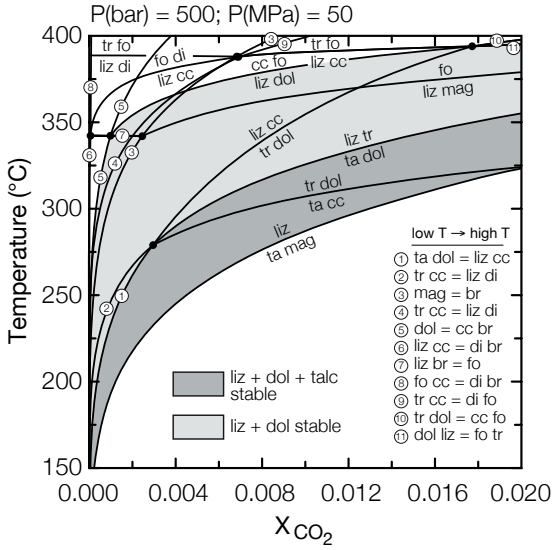

Figure 7: T–X_{CO_2} diagram of stable phases for the system CaO–MgO–SiO_2–H_2O–CO_2. Calculations assume saturation with a H_2O–CO_2 fluid using the Perple_X application suite (Connolly, 2005) using the Holland and Powell (2004) thermodynamic database (hp02ver.dat) modified by Connolly and Kerrick (2002). *Abbreviations*: *br*=brucite, *cc*=calcite, *di*=diopside, *dol*=dolomite, *fo*=forsterite, *liz*=lizardite, *ta*=talc, *tr*=tremolite. Al-free chlorite (afchl), which has the same mineral structure and composition as Al-free lizardite, was used as "lizardite" in the calculations. Reactions with wollastonite and quartz are also stable, but do not bear on this discussion and were left off the diagram to improve clarity.

Alteration model

It is useful to recognise two stages in the alteration history of Snap Lake. The first stage takes place during emplacement and cooling of the intrusion. Its duration will be short and is controlled by heat loss, crystallisation and penetration of ground water into the intrusion, resulting in the onset of alteration. For an intrusion 4 m thick, the conductive cooling time is estimated as less than 60 days using a thermal diffusivity of 8×10^{-7} m^2s^{-1} (Gernon et al., 2012). The second stage is after the cooling of the intrusion is complete—and in principle could be 537 million years (i.e. since emplacement; Agashev et al., 2008)—although we think it likely that alteration was complete in much less time than this. Summarising the above evidence and arguments, deuteric alteration can be ruled out and the alteration of the kimberlite and granitoid host rocks can be explained by serpentinising ground waters.

Severe alteration is the most characteristic features of kimberlites and overprints many of the magmatic stage features of kimberlites. In order to infer magmatic history, the alteration history must first be unraveled. Fulop et al. (2018) start from the assumption that many textures and mineral assemblages are magmatic in origin unless proven otherwise. In other areas of petrology the approach is that textures and features in highly altered rock are the consequence of alteration unless proven otherwise.

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